Ice Cloud Optical Thickness and Extinction Estimates from Radar Measurements

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ABSTRACT

A remote sensing method is proposed to derive vertical profiles of the visible extinction coefficients in ice clouds from measurements of the radar reflectivity and Doppler velocity taken by a vertically pointing 35-GHz cloud radar. The extinction coefficient and its vertical integral, optical thickness τ, are among the fundamental cloud optical parameters that, to a large extent, determine the radiative impact of clouds. The results obtained with this method could be used as input for different climate and radiation models and for comparisons with parameterizations that relate cloud microphysical parameters and optical properties. An important advantage of the proposed method is its potential applicability to multicloud situations and mixed-phase conditions. In the latter case, it might be able to provide the information on the ice component of mixed-phase clouds if the radar moments are dominated by this component. The uncertainties of radar-based retrievals of cloud visible optical thickness are estimated by comparing retrieval results with optical thicknesses obtained independently from radiometric measurements during the yearlong Surface Heat Budget of the Arctic Ocean (SHEBA) field experiment. The radiometric measurements provide a robust way to estimate τ but are applicable only to optically thin ice clouds without intervening liquid layers. The comparisons of cloud optical thicknesses retrieved from radar and from radiometer measurements indicate an uncertainty of about 77% and a bias of about −14% in the radar estimates of τ relative to radiometric retrievals. One possible explanation of the negative bias is an inherently low sensitivity of radar measurements to smaller cloud particles that still contribute noticeably to the cloud extinction. This estimate of the uncertainty is in line with simple theoretical considerations, and the associated retrieval accuracy should be considered good for a nonoptical instrument, such as radar. This paper also presents relations between radar-derived characteristic cloud particle sizes and effective sizes used in models. An average relation among τ, cloud ice water path, and the layer mean value of cloud particle characteristic size is also given. This relation is found to be in good agreement with in situ measurements. Despite a high uncertainty of radar estimates of extinction, this method is useful for many clouds where optical measurements are not available because of cloud multilayering or opaqueness.

1. Introduction

Numerous studies have confirmed the fundamental role of clouds in the earth's radiation budget and climate system (e.g., Wielicki et al. 1995). Ice clouds have been identified as one of the most uncertain components of this system (e.g., Stephens et al. 1990; Tian and Ramanathan 2002). These clouds are often several kilometers thick (Kosarev and Mazin 1991), but in many cases they are optically "nonblack" with visible optical thicknesses less than about 4. An adequate quantitative description of these clouds in models is very important because they regularly cover up to 40% of the globe (Raschke 1993). Ice clouds influence the radiation field of the earth–atmosphere system in both solar (shortwave) and thermal IR (longwave) bands. The relative significance of the ice cloud role in these two bands

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(i.e., the so-called greenhouse-vs-albedo effect) leads to differing atmospheric cooling and heating rates, as well as to different impacts on surface and top of the atmosphere energy budgets.

Cloud radiative properties are determined by their macrophysical parameters, such as geometrical thickness, altitude, layering, and horizontal extent, together with their microphysical parameters. The radiatively important microphysical properties of ice clouds include ice water content (IWC), characteristic cloud particle size (e.g., effective radius, median, or mean sizes), and also particle shape (habit). The wide interest in clouds' role in the climate system has stimulated development of satellite- and ground-based remote sensing methods to retrieve these radiatively important microphysical properties. For cloud retrievals, satellite methods usually rely on passive measurements provided by different visible and infrared channels of instruments, such as the Moderate Resolution Imaging Spectroradiometer (MODIS), Advanced Very High Resolution Radiometer (AVHRR), Visible Infrared Scanner (VIRS), and Clouds and the Earth's Radiant Energy System (CERES). The passive measurements are able to provide only vertically integrated or layer mean parameters and generally are not suitable for multilayer cloud scenes. However, for the adequate cloud representation in models, it is necessary to know vertical profiles of the cloud radiative properties such as heating/cooling rates and radiative fluxes.

Most ground-based cloud retrieval methods employ suites of different vertically pointed remote sensors. Cloud radars commonly operating at Ka (at frequencies around 35 GHz) or W (at frequencies around 90 GHz) bands are often a centerpiece of these suites. A combination of a cloud radar and an IR spectrometer/radiometer allows retrievals of layer mean values of IWC and the cloud particle characteristic size (Matrosov et al. 1992; Mace et al. 1998) or vertical profiles of IWC and the characteristic size (Matrosov 1999). Combinations of lidar and radar measurements are also used for retrievals of microphysical profiles (Donovan and van Lammeren 2001; Wang and Sassen 2001), though the attenuation of lidar signals causes limitations of the applicability of lidar–radar approaches.

Recently, new ice cloud microphysical methods based on Doppler radar–only measurements have been suggested (Matrosov and Heymsfield 2000; Matrosov et al. 2002; Mace et al. 2002). These methods use vertical profiles of radar reflectivity and Doppler velocity to retrieve profiles of IWC and the cloud particle characteristic size. As compared with the methods that use combinations of microwave (e.g., radar) and optical (e.g., lidar or/and IR radiometer) instruments, the applicability of these new methods is extended to multilayer and optically thick clouds. Problems associated with a mismatch of fields of view of different instruments are nonexistent for the radar-only approaches. On the other hand, the use of Doppler information requires time averaging, and so the effective time resolution of these methods is relatively low (e.g., 20 min or so). Radars also often do not detect clouds or portions of clouds with populations of very small particles. This limitation is dependent on radar sensitivity and the implications of the radar missing small particles changes, depending on the applications.

Accurate modeling of the ice cloud impact on the earth's radiation budget and climate requires parameterization of cloud optical properties, such as the extinction coefficient, optical thickness, phase function, and single-scattering albedo, in terms of retrieved or assumed cloud microphysical parameters. A number of parameterization schemes have been suggested in the last few years (e.g., Ebert and Curry 1992; Fu 1996; Yang and Liou 1998; Key et al. 2002). Some of the cloud optical properties, however, can be retrieved directly from remote sensing measurements. The retrieved optical properties can then be incorporated directly into the radiation schemes of different models. Long-term remote sensing retrievals of cloud optical properties also permit a straightforward assessment of the frequency distributions and other important statistics of these cloud properties.

An extension of the Doppler radar–only ice cloud microphysical retrieval method, suggested recently by Matrosov et al. (2002), also allows retrievals of the vertical profiles of the cloud extinction coefficient. Radars operate at a wavelength that is vastly different from optical wavelengths and are not usually associated with any kind of optical measurements. Therefore, for the purpose of practical use, it is necessary to understand and quantify the accuracy of the optical product obtained from radar measurements. Although direct measurements of cloud-specific extinction is not readily available, measurements of its integral (i.e., the cloud optical thickness) can be made in a rather straightforward way using optical measurements. This study compares radar- and optically (i.e., radiometrically) derived optical thicknesses of ice clouds observed during the year-long Surface Heat Budget of the Arctic Ocean (SHEBA) experiment (Uttal et al. 2002) and estimates the accuracy of radar retrievals.

2. Radar retrievals of the cloud extinction coefficient and optical thickness

The Doppler radar–only ice cloud microphysical retrieval method (Matrosov et al. 2002) uses profiles of Doppler velocity and radar reflectivity to retrieve profiles of IWC (i.e., cloud ice mass) and median volume particle size (D_0), which describes the whole particle size distribution. The individual particle sizes here are understood in a manner as they are usually measured by aircraft probes. Often these sizes are measured as the mean of maximum chord lengths measured in the parallel and perpendicular directions relative to the probe photodiode array (Korolev and Strapp 2002). Cloud visible extinction is proportional to the particle geometrical cross-sectional area. This fact, and the existence of general empirical relations between individual particle size (D), cross-sectional area (A), and mass (m), allow the use of a similar Doppler radar–only technique for retrievals of the cloud extinction coefficient (α) and D_0. As outlined by Matrosov et al. (2002) this technique is based on the expression for the radar reflectivity Z_e as a function of D_0 and α,

$$Z_e = X\alpha D_0^4, \tag{1}$$

where X is a dimensionless proportionality factor that depends on D_0 if the ice particle bulk density ρ changes with D. This factor depends also on particle shape and details of the particle size distribution. It has been suggested (e.g., Mace et al. 2002) than an exponential distribution usually satisfactorily describes ice cloud particle spectra for larger particles that are best seen by radar. The variability of the factor X when the size distribution function is described by the zeroth- (i.e., exponential), first-, or second-order gamma function is quite modest (Matrosov at al. 2002), and so, for the purpose of this work, it is assumed that this distribution is exponential. The potential multimodality of the size distribution would complicate the relation between X and D_0. This would result in higher errors of extinction retrievals from radar measurements. The relatively high retrieval uncertainties discussed in sections 5 and 6 are, in part, due to the approximate nature of the size distribution assumption.

It is assumed that the visible extinction cross section of a particle is equal to 2A. Heymsfield et al. (2002) found the following m–A–D relation from in situ measurements of ice cloud particles:

$$mA^{-1} \approx 0.038D^{0.576} \text{ (cgs)}. \tag{2}$$

Though this relation was obtained from the in situ samples of midlatitude winter ice clouds, it is used here because there is no such relation available for the Arctic clouds. Some justification for this use is that the particle habits found in the midlatitude winter ice clouds and the Arctic clouds are similar. Typically particles have irregular shapes with a very low fraction ($<$8%) of pristine crystals (Korolev et al. 1999). It should be admitted, however, that some uncertainty in relation (2) exists, though, as will be shown below, the use of (2) results in general correspondence between optical thickness values derived from radar measurements and independently from radiometer measurements.

Using (2) and the particle bulk density assumption, one can obtain approximations for the factor X (Matrosov et al. 2002). The Locatelli and Hobbs (1974) particle mass–size relation leads to the following bulk density assumption:

$$\rho \text{ (g cm}^{-3}) \approx 0.07D^{-1.1} \text{ (mm)}. \tag{3}$$

The solid ice density (0.9 g cm^{-3}) is used when (3) provides values greater than 0.9 g cm^{-3}. The assumption (3) was shown to be generally appropriate for cirrus (Brown and Francis 1995), though some later microphysical studies including Arctic clouds (e.g., Korolev and Strapp 2002) indicate that it might somewhat overestimate particle masses. Assuming the spherical cross sections for smaller particles when (3) provides values of A that exceed those for spheres, the following approximation for the coefficient X can be obtained:

$$X \approx \begin{cases} 4.6 \times 10^{-4} D_0^{-1.8} & (D_0 > 58\ \mu\text{m}), \\ 3 \times 10^{-7} & (D_0 \lesssim 58\ \mu\text{m}). \end{cases} \tag{4}$$

Substituting (4) in (1) will result in Z_e (mm^6 m^{-3}), α (m^{-1}), and D_0 (μm). For typical values of particle median sizes ($D_0 < 300\ \mu$m), the approximation (4) is generally within 40% of the one obtained by Matrosov et al. (2002), using an older m–A–D assumption.

Estimates of α from Doppler radar measurements are done in two steps. First, the profile of D_0 is obtained from the time-averaged vertical profile of Doppler velocity measurements, using the approach presented by Matrosov et al. (2002), and then the extinction coefficient profile is calculated from (1) using the profile of the radar reflectivity. A typical averaging time is about 20 min. As a result of such averaging residual vertical air motion is usually significantly less than the reflectivity-weighted particle fall velocities. However, clouds with strong vertical air updrafts or downdrafts may require longer averaging times (Matrosov et al. 2002), and, in some extreme situations, Doppler velocity averaging would not produce satisfactory estimates of cloud particle terminal fall velocities (i.e., D_0 profiles).

Figure 1 shows an example of retrievals for one of the ice clouds observed during the SHEBA field experiment. A geometrically thick ice cloud was observed over a period of more than 18 h on 28–29 April 1998. Twenty-minute averages of the radar reflectivity and Doppler velocity observed in this cloud by the 8-mm-wavelength cloud radar (MMCR) are depicted in Figs. 1a and Fig. 1b. Figures 1c and 1d show the retrievals of the extinction coefficient and particle median volume size, D_0 (in terms of ''maximum chord length'' equivalent spherical diameters). Microphysical retrievals for this case were previously shown to compare well with aircraft measurements (Matrosov et al. 2002).

Uncertainties in the different retrieval assumptions coupled with the uncertainties of radar measurements and estimates of D_0 from the Doppler velocity data could result in about a factor of 2 (or even more) uncertainty (i.e., $+100$, -50%) in the extinction coefficient retrievals (Matrosov et al. 2002). These theoretical estimates of retrieval uncertainties might seem rather large but the radar assessment of the cloud extinction is still useful because it can be obtained for multilayer clouds when optical sensors are either blocked by the lower liquid cloud layers or their signals are severely attenuated. Passive satellite measurements cannot properly handle multilevel clouds either. Another important advantage of radar measurements is that they may be

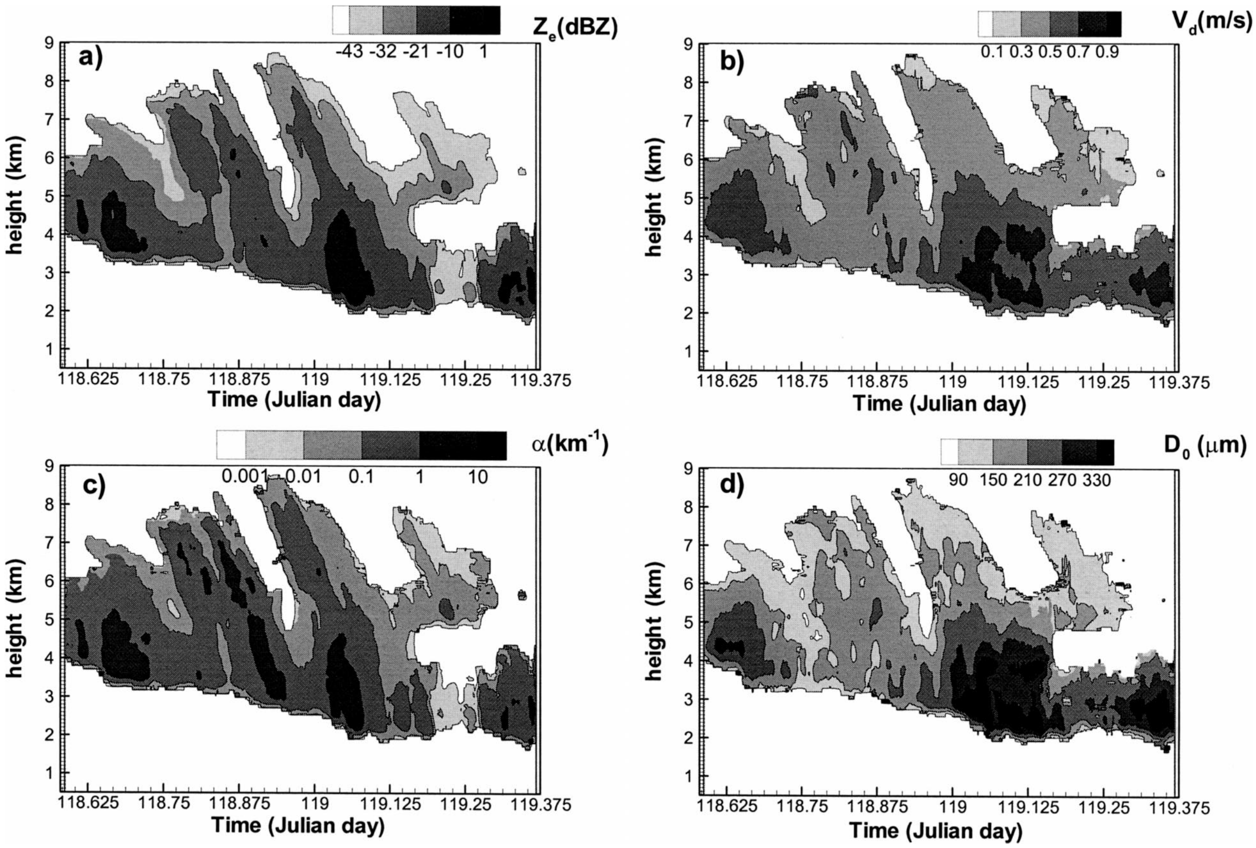

FIG. 1. Twenty-minute averages of measurements of (a) radar reflectivity and (b) Doppler velocity, (c) retrieved extinction coefficients, and (d) particle mean sizes in an ice cloud observed on 28–29 Apr 1998.

able to provide the extinction of the ice component of the nonprecipitating, mixed-phase clouds if contributions of ice particles to Doppler radar moments are much larger than contributions by water drops that are often much smaller than ice particles.

Uncertainties of the radar estimates of cloud optical thickness, τ, which is the vertical integral of α, are likely to be smaller than those for the extinction coefficient, because in the course of integration there is a partial compensation of errors in α that have opposite signs. The quality of optical parameter retrievals using radar measurements, however, can also be assessed by direct comparisons with results of measurements taken by optical instruments. Such comparisons can only be performed for single-layer ice clouds that are unobstructed as ''seen'' by such instruments. The SHEBA dataset provided a convenient opportunity for such comparisons.

Remote sensing retrievals of cloud extinction profiles from lidar measurements can be done for relatively thin clouds. These retrievals are possible with the use of specialized lidars, such as the Raman lidar or the University of Wisconsin high-spectral-resolution lidar (HSRL), when the aerosol (cloud) extinction profile can be directly obtained (Piironen and Eloranta 1994). During SHEBA such lidars were not available. The backscatter data from the available depolarization and backscatter unattended lidar (DABUL) have not been absolutely calibrated. Optical thickness estimates, however, could be performed in a relatively straightforward way using the radiometric measurements. Note also that the radiometric approach to estimate cloud optical thickness has an advantage over the lidar-based approaches in the sense that it can be applied to optically thicker ice clouds because, for typical particle sizes, the extinction optical thickness is about 2 times the absorption optical thickness.

3. Estimations of cloud optical thickness from AERI

Measurements of the atmospheric emitted radiance interferometer (AERI) provided a means to estimate cloud absorption optical thickness τ_a at the SHEBA ice station. This instrument provided high-resolution (1 cm^{-1}) spectral measurements of the downwelling radiance in a range from 500 (20 μm) to 3300 (3 μm) cm^{-1} every 10 min. The AERI data in the IR ''window'' centered at 900 cm^{-1} ($\pm$12.5 cm^{-1}) were used for calculating τ_a based on the following expression for the brightness temperature of the downwelling radiation T_b (Matrosov et al. 1998):

$$B(T_b) = B(T_e)[1 - \exp(-\tau_a) + \delta\epsilon]P_a + (1 - P_a)B(T_a) + R_g, \qquad (5)$$

where B is the Planck function for 900 cm^{-1}, P_a is the transmittance of the atmosphere between the cloud and the ground, $\delta\epsilon$ is the term correcting cloud emissivity for scattering effects, and T_e and T_a are the cloud- and the atmosphere-emitting temperatures. The first term in (5) describes the radiation of the cloud attenuated by the atmospheric layer, the second term shows the radiation contribution from this layer, and R_g accounts for the ground radiation reflected by the cloud. The ground term R_g does not contribute significantly to the total downwelling radiation (Matrosov et al. 1998), and it is usually neglected.

Matrosov et al. (1998) suggested a procedure to find T_e and T_a from vertical temperature profiles (known from a nearby radiosonde sounding) and radar-measured cloud boundaries. Note that T_e differs from the midcloud temperature and it is approximately equal to the thermodynamic temperature at a certain level inside the cloud. The ratio of the cloud absorption optical thickness from the cloud base to this level, τ_0, and the total absorption optical thickness, τ_a, depends on τ_a and can be approximated for $\tau_a < 6$ as $\tau_0/\tau_a \approx (0.006\tau_a^2 - 0.09\tau_a + 0.5)$ (Matrosov et al. 1998). For the optically thickest ice clouds, where the radiometric assessment of τ_a is still possible, the effective emitting cloud temperature T_e is approximately equal to the thermodynamic temperature at a cloud level where the absorption optical thickness from the cloud base is about 1. The effective atmosphere-emitting temperature T_a is mostly determined by the vertical profile of the water vapor. For typical water vapor profiles, T_a is approximately equal to the thermodynamic temperature at an altitude of about 1 km above the ground.

Atmospheric transmittance P_a is determined mostly by the water vapor. Because water vapor tends to be concentered in the lower atmosphere, it is assumed that most of the total vertically integrated water vapor amount (WVA) is in the layer between the cloud and the ground. A simplified relation suggested by Matrosov et al. (1998) for midlatitude conditions,

$$P_a \approx 1.02 \exp(-0.103\mathrm{WVA}) \quad (\mathrm{WVA\ in\ centimeters}), \qquad (6)$$

was found also to be valid for the relatively dry conditions that usually are present in the Arctic. This expression was verified by comparing the cloud-free 900 cm^{-1} AERI measurements with the prediction from (5), where only the second term is retained and WVA values are obtained from the dual-channel microwave radiometer (31.8 and 23.4 GHz) measurements or from radiosonde soundings. The differences between calculated and measured brightness temperatures were, as a rule, within a few kelvins.

The emissivity correction term $\delta\epsilon$ is defined as the difference between cloud emissivity when accounting for scattering and the cloud emissivity under the assumption of a ‘‘pure absorption’’ regime. The scattering effects are twofold. On one hand, scattering increases the downward radiation by redirecting photons from other directions. On the other hand, it removes photons from the downward direction through volume scattering. The first effect enhances the downwelling radiation and the second effect reduces it. These two effects tend to approximately balance each other in the thermal IR radiation region. Often, scattering effects are neglected altogether (i.e., Inoue 1985). Modeling of the term $\delta\epsilon$ was performed using the discrete ordinate algorithm for radiative transfer calculation (Stamnes et al. 1988).

The calculations were performed using the 48-stream algorithm version for three ice cloud particle models with equal volume spherical diameters of 22, 55, and 190 μm, which approximately correspond to the small (C20), medium (CS), and large (CU) ice particle clouds, according to the nomenclature presented by Minnis et al. (1993). The spherical model is considered here because ice particles sampled in Arctic ice clouds are mostly irregular particles and the fraction of pristine single hexagonal crystal is very small (Korolev et al. 1999).

Figure 2 shows the correction $\delta\epsilon$ as a function of the total (extinction) cloud optical thickness, τ. This correction is the largest for the smallest particle model (C20). The particle size dependence of this correction is approximately inversely proportional to the square root of particle size. An overall approximation of $\delta\epsilon(\tau)$ can be achieved by a cubic polynomial function as shown in Fig. 2. In the Environmental Technology Laboratory (ETL)’s retrieval procedure, the term $\delta\epsilon$ is estimated based on the layer mean value of the cloud particle size obtained from Doppler velocity measurements (Matrosov et al. 2002). Note, however, that this correction is rather small, and the pure absorption assumption is generally a good approximation for retrieving cloud optical thickness in the thermal IR range where ice absorption is significant.

The radiometric approach described above provides the absorption optical thickness τ_a. It is assumed that τ_a and the total visible extinction optical thickness τ are related as $\tau \approx 2\tau_a$. This assumption has been validated by experimental comparisons presented by Matrosov et al. (1998) and by theoretical considerations for the large (in comparison with the wavelength) particle regime (i.e., the geometrical optics regime) where extinction efficiency is approaching 2 and the absorption efficiency is approaching 1. Note that some model studies using the hexagonal ice crystals resulted in a factor of 2.1 for the ratio τ/τ_a (Fu and Liou 1993), and some other observational estimates (Platt et al. 1987) showed this ratio in the range of values from 1.8 to 2.3. The variability

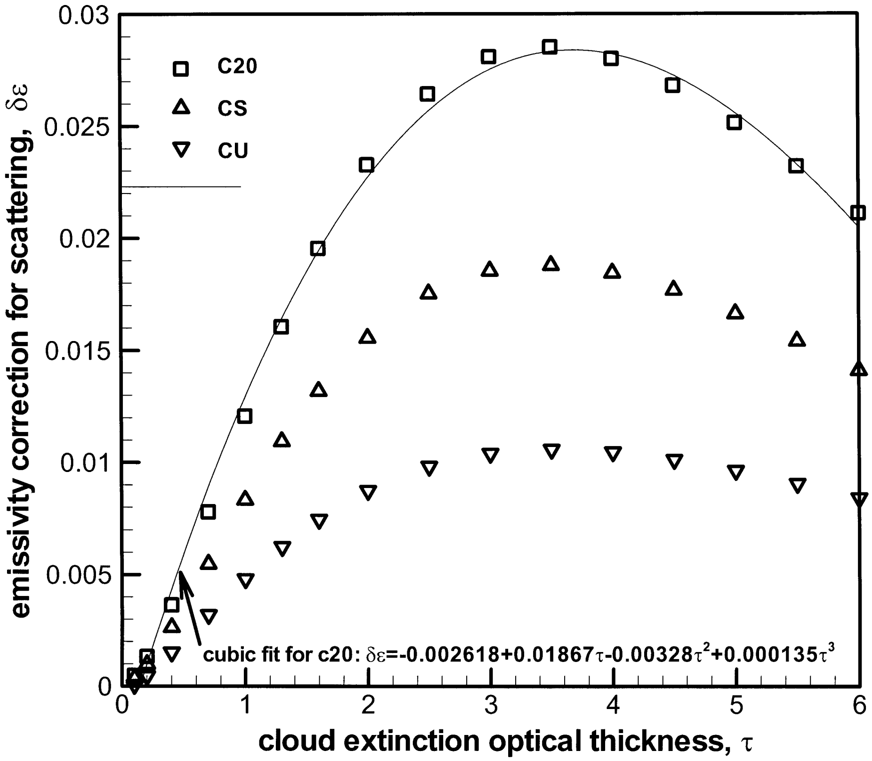

FIG. 2. Emissivity correction for scattering as a function of the cloud extinction optical thickness for the different models (C20, CS, and CU) of cloud particles.

of this ratio can be explained, in part, by the fact that the smallest cloud particles are not in the geometrical optics regime for the thermal infrared wavelengths. Such particles may actually be missed by the radar because of radar sensitivity limitations. Given inevitable retrievals errors, a value of 2 is assumed for this ratio for the purpose of this work.

Practically, radiometric estimations of cloud optical thickness are possible for unobstructed ice cloud layers with τ_a less than about 3, which corresponds to a cloud emissivity of about 0.95. For larger values of absorption optical thickness clouds become essentially ''black,'' and the radiation saturation effects prevent retrievals. Note that lidar signals can become practically extinct in clouds already at the level of about $\tau_a \approx 1.5$, which corresponds to the round-trip attenuation of about $\exp(-2\tau) = \exp(-6)$, assuming $\tau = 2\tau_a$. This illustrates the fact that the radiometric approach for estimating cloud optical thicknesses is applicable to a wider range of ice clouds than any method that uses lidar measurements. It should be noted, however, that for larger particles and fields of view, multiple scattering and other factors may lead to an increased penetration of lidar signals. During SHEBA, the radiometric approach was applied to the ice clouds with no intervening liquid layers (as identified by the dual-channel microwave radiometer measurements). The AERI data were available only for the first 8 months (November 1997–June 1998) out of the yearlong SHEBA period (Uttal et al. 2002).

4. Comparisons of radar- and radiometer-derived optical thicknesses

The radiometer approach to derive optical thickness described in section 3 uses some radar-based information, such as the location of cloud boundaries for cloud temperature estimates, and radar-derived particle size information for estimates of the scattering correction $\delta\epsilon$. This information, however, presents relatively minor adjustments to purely radiometrical estimates, and the optical thickness values derived using the radar and radiometer (i.e., AERI and microwave radiometer) data are essentially independent. Hereinafter, optical thicknesses derived using the method described in section 2 are referred to as τ from radar data and results of the approach from section 3 are referred to as τ from radiometric data.

The radiometric approach (section 3) presents a relatively straightforward and robust way to derive cloud optical thickness. The radar method (section 2) is a novel approach that suggests using a nonoptical instrument to retrieve cloud optical parameters. Although possible retrieval errors of the radar method were estimated (Matrosov et al. 2002), direct comparisons with optical (ra-

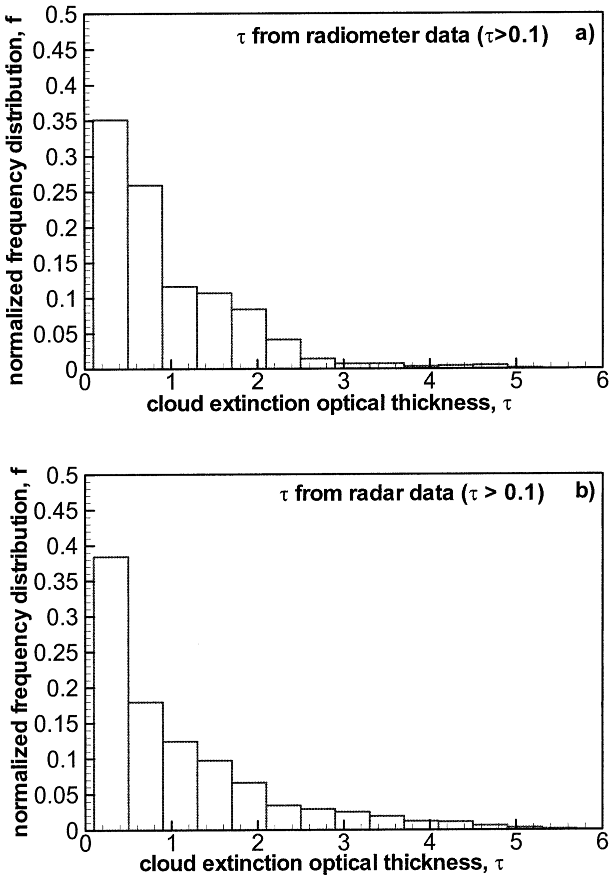

FIG. 3. Histograms of the normalized frequency distributions of ice cloud optical thickness from (a) radiometric retrievals and (b) radar-only retrievals.

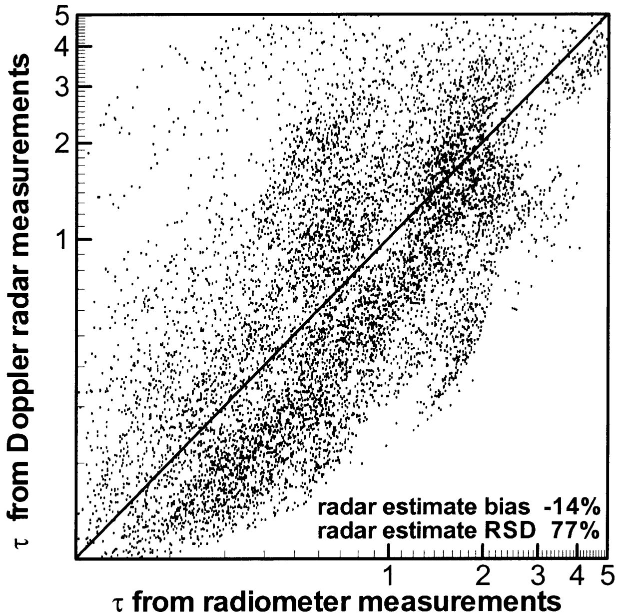

FIG. 4. A scatterplot of cloud optical thickness derived from radiometric and the radar-only measurements. Radar-only values were obtained by integrating extinction coefficients estimated using (1). The bias and the RSD of radar estimates are also shown.

diometric) estimates of τ from radiometer data will provide a better insight into the potential usefulness of the radar method. Relatively long-term SHEBA observations also provide possibilities of statistical analysis of radar retrievals of τ.

Figure 3 shows histograms of the normalized frequency distributions of the cloud optical thicknesses derived from collocated radar and radiometric measurements made with vertically pointed instruments during the SHEBA period. Very tenuous ice clouds with $\tau < 0.1$ were excluded from the further analysis because retrieval errors for such clouds are rather high (greater than 100%). The frequency distributions in Figs. 3a and 3b are similar, though there are some minor differences indicating that, for smaller values of τ, the radar results may be negatively biased in comparison with radiometric data. Generally, both approaches show a gradual decrease in the occurrence frequency as the cloud optical thickness increases. The data are presented for clouds when both retrievals were available (i.e., for ice clouds that were not obstructed by liquid).

A scatterplot of cloud optical thicknesses retrieved from radiometric and radar measurements is shown in Fig. 4. The correlation coefficient characterizing data scatter is about 0.62. The mean bias of the radar data from the radiometric estimates is about -14%, and the corresponding relative standard deviation (RSD) is about 77%. The RSD characterizes the total uncertainty of radar retrievals of cloud optical thickness relative to the radiometrically derived values. Figure 5 shows the biases and standard deviations of radar estimates as a function of cloud optical thickness τ. It can be seen that uncertainties of radar retrievals generally decrease with τ. It is likely that the relative contributions of small particles to the total optical thickness are larger for smaller values of τ, which results in less accurate radar estimates of τ because of the radar-poor sensitivity to small particle populations.

The fact that the data scatter in Fig. 4 (and Fig. 6) is significant does not disqualify the radar approach for estimating τ. Indeed, the RSD value of 77% represents an average retrieval uncertainty to be about a factor of 2 (i.e., $+100\%$, -50%). Similar or slightly smaller uncertainties are associated with remote (and also in situ) estimates of such cloud microphysical parameters as ice water content (Matrosov et al. 2002). The radar-based retrievals of cloud optical thickness have an advantage over optical measurements because they are applicable to multilayer cloud scenes, which are common in cloud remote sensing. Once the associated uncertainties are realized, optical thickness estimates provided by radar can be a very useful input for different models.

Measured radar moments are weighted by the product $\rho^2 D^6$ (Matrosov and Heymsfield 2000), where ρ is the bulk density and D is the particle size. For smaller particles with a nearly solid ice density, it results in weight-

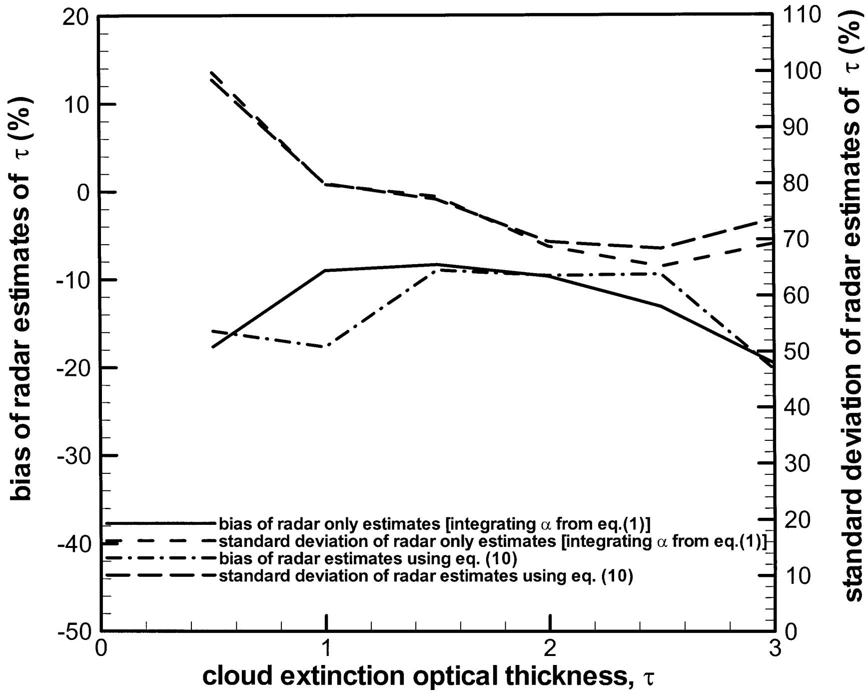

FIG. 5. Biases and standard deviations of optical thickness radar estimates as a function of τ.

ing by the 6th moment of the particle size distribution (PSD). For large particles, this product is approximately proportional to the 3.8th moment of the PSD because the bulk density of such particles is proportional to $D^{-1.1}$, according to the Locatelli and Hobbs (1974) relation mentioned above. There is a transition zone where the proportionality changes from the 6th moment to the 3.8th moment. Thus, larger cloud particles dominate radar signals. The smaller particles are accounted for by the assumption of the shape of the PSD. However, the smallest particle populations might not be adequately described by the exponential PSD assumed here. Extinction, on the other hand, is much more sensitive to smaller particles than radar moments. The role of small particles is greater for cloud extinction/optical thickness than for cloud ice water content/ice water path (IWP). An inadequate representation of small particles in radar-based retrievals is one possible explanation of the negative bias of radar retrievals of τ in comparison with radiometric estimates. Another possible explanation is the uncertainty of the coefficients in the density size and *m–A–D* relations, which provide a connection between microphysical and optical (i.e., extinction) properties. Potentially, it is possible to ‘‘tune’’ these coefficients to eliminate the existing bias. For example, tuning the value of the coefficient (3a) could result in the unbiased radar estimates of τ. However, the existing bias is not very significant given the expected overall retrieval uncertainties, and we will leave a subject of tuning *m–A–D* relations for further studies when more data become available.

5. Relations between cloud microphysical and optical properties

In model parameterizations of radiative properties of ice clouds, the extinction coefficient α is usually expressed in terms of IWC and some characteristic particle size D_c. Virtually all of the parameterizations imply the proportionality between α and IWC. As for the dependence on particle size, different parameterizations assume either simple inverse proportionality of α and D_c,

$$\alpha = \mathrm{IWC} b_0 D_c^{-1}, \tag{7}$$

or a general linear relation between α and D_c^{-1} (e.g., Fu and Liou 1993),

$$\alpha = \mathrm{IWC}(a + b_1 D_c^{-1}), \tag{8}$$

where b_0, b_1, and a are the parameterization coefficients. Different characteristic cloud particle sizes can be used in such parameterizations. One choice is the effective size D_e, which, by some definitions, is proportional to the ratio of the third and second moments of the PSD. For a given assumption of the PSD shape (e.g., exponential) and particle bulk density, D_e corresponds in a one-to-one manner to other characteristic sizes that describe the whole PSD. Those sizes include median volume size D_0, median mass size D_m, and mean size D_{mean}. Even if the particles are assumed to be nonspherical, it is convenient to express these sizes in terms of diameters of the corresponding spheres.

The parameterization (7) is similar to that of water clouds and it is appropriate when the particle bulk density is constant. For the size-dependent particle bulk density, which is the case for ice clouds, the parameterization (8) is more appropriate. The relation for the cloud optical thickness can be obtained by vertically integrating (8),

$$\tau = \mathrm{IWP}(a + b\overline{D}_c^{-1}), \tag{9}$$

where IWP is ice water path, and the characteristic size $\overline{D}_c$ represents the whole ice cloud layer.

Matrosov and Snider (1995) used a relation similar to (9) but for the absorption optical thickness τ_a. Based on the analysis of remote sensing data for many one-layer ice cloud observational cases, they found that a' varies approximately between 0.008 and 0.012 and b' varies between about 1.6 and 2.2 if the IWC-weighted vertical average of median mass particle sizes expressed in micrometers is used as $\overline{D}_c$, and the units of IWP are in grams per meter squared. The superscripts (′) in a' and b' make a distinction that they refer to τ_a and not τ in (9). With the assumption that $\tau = 2\tau_a$ and using $\overline{D}_0$ as $\overline{D}_c$, the variability of the parameterization coefficients in (9) is 0.016–0.024 for a and 3.6–4.9 for b. Mean values of these coefficients are 0.02 and 4.2, respectively. It was assumed here that the median mass particle size is on average 10%–20% smaller than the median volume particle size (Matrosov et al. 1998). For multilayer ice clouds, $\overline{D}_c$ represents an average value through all of the ice cloud layers.

Note that according to (9) for the population of small particles when the second term in (9) is dominant, optical thickness is approximately inversely proportional to characteristic particle size. For population of larger particles, the dependence of τ on particle size becomes progressively smaller and a value of IWP approximately determines cloud optical thickness. This is because the particle size–bulk density dependence results in both IWP and τ becoming approximately proportional to the same moment of the PSD for larger particles.

Because the Doppler radar–based method (Matrosov et al. 2002) allows retrievals of vertical profiles of IWC and D_0, predictions of optical thickness using (9) can be compared with the radiometric measurements of τ. This will represent another comparison of independently estimated optical thicknesses. Figure 6 shows the scatterplot of the radiometrically derived optical thickness and the results obtained from microphysical retrievals using

$$\tau = \mathrm{IWP}(0.02 + 4.2\overline{D}_0^{-1}),$$
$$(\overline{D}_0 \text{ in micrometers; IWP in grams per meter squared}). \tag{10}$$

It can be seen from comparing Figs. 4 and 6 that the data scatter is similar. The bias of the optical thickness estimates from the cloud microphysics is −17% and the RSD is about 79%, which is slightly greater than from direct extinction estimates from radar data. The overall

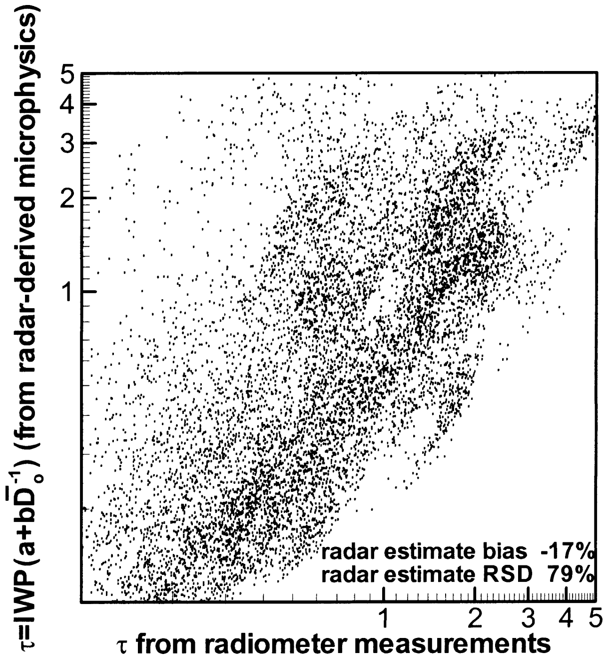

FIG. 6. Same as Fig. 4, but the radar–only optical thicknesses are obtained from the radar-based microphysical retrievals using (10) and cloud microphysical parameters derived using the method described by Matrosov et al. (2002).

closeness of the results in Figs. 4 and 6 can be explained by the similarity of the approaches and assumptions used to retrieve cloud extinction as described in section 2, and cloud microphysical properties as described by Matrosov et al. (2002). Both approaches use the same input information, that is, the vertical profiles of the radar reflectivity and the Doppler velocity. This means that cloud optical thickness estimates can be obtained from Doppler radar measurements directly or from previously retrieved microphysical parameters using (10). An advantage of the direct approach, which is presented in section 2, is that it also provides a relatively straightforward way of estimating extinction profiles. Another advantage of the direct radar–only approach is that it handles multilayer clouds better. Note that (8), with coefficients from (9), can also be used for obtaining extinction profiles from known microphysical profiles (i.e., IWC and D_0).

Data collected from several balloonborne replicators in midlatitude cirrus clouds were used to verify the relations between cloud optical thickness and IWP. The replicator (Miloshevich and Heymsfield 1997) provides high-quality PSD measurements with a resolution of about 2 μm from about 10 μm to 2 mm in particle size. In comparison with the traditional two-dimensional cloud (2D-C) probes, replicators provide much better measurements for smaller particles. In situ estimates of the cloud extinction coefficient α are calculated using replicator data on the particle cross-sectional area A assuming an extinction efficiency of 2, and the particle

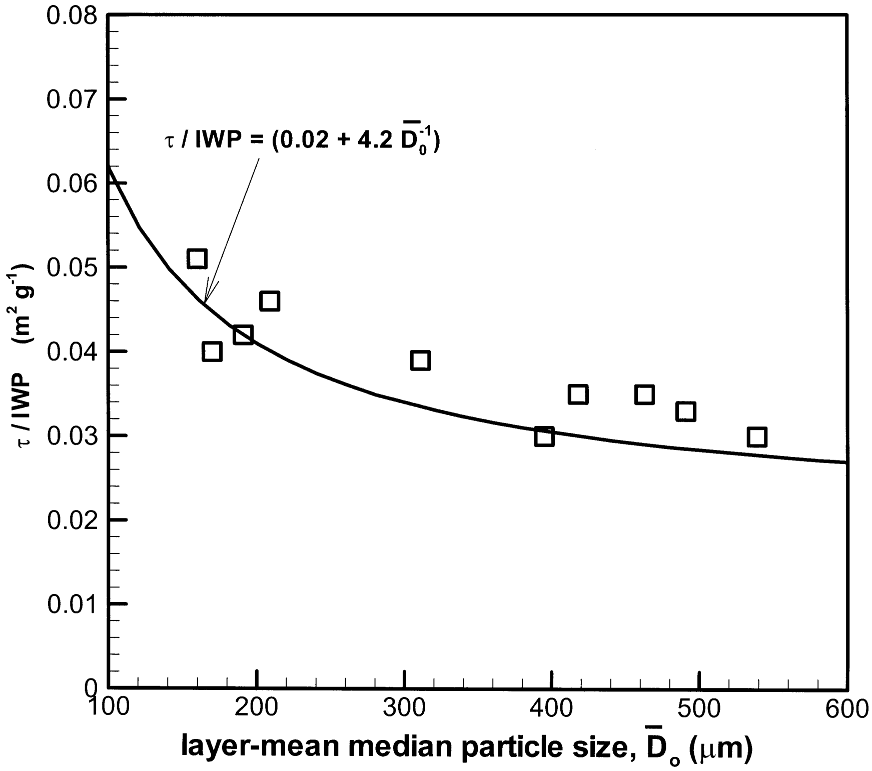

FIG. 7. The τ/IWP ratio as a function of layer mean median particle size as derived from in situ measurements.

mass (i.e., IWC) is calculated using the habit-dependent mass–size relations. The details of such calculations are given by Heymsfield et al. (2002). The optical thickness and IWP values are then obtained by integrating α and IWC values.

Figure 7 shows τ/IWP ratios from replicator data as a function of the mean layer value of particle median size $\overline{D}_0$. The data were obtained from several Lagrangian aircraft spiral descents and replicator-balloon ascents during the First International Satellite Cloud Climatology Project (ISCCP) Research Experiment (FIRE-II) at Coffeyville, Kansas; and several Atmospheric Radiation Program (ARM) intensive observation periods (IOPs) near Lamont, Oklahoma (Heymsfield and Miloshevich 2003). Though the Lagrangian data do not provide optical thickness and IWP estimates in the vertical direction, they still present a valid comparison for the relation (10) because all of the profiles are considered along the ascent line and no direct retrievals from vertically pointed radar data are involved in this comparison.

It can be seen that there is generally good agreement between τ/IWP ratio values from in situ measurements and those predicted by the relation (10). There is a distinct flattening of the τ/IWP ratio dependence on the layer mean particle size for larger values of $\overline{D}_0$, though values from (9) are a little lower than in situ data. On the smaller particle end, that the agreement is better and a clear trend for the in situ τ/IWP ratios to increase as $\overline{D}_0$ diminishes is obvious. Though qualitative agreement is good, more data are obviously necessary to get more reliable quantitative comparisons.

6. Relations between radar-derived and optical cloud particle characteristic sizes

The Doppler radar retrieval method described here provides estimates of median volume particle size $\overline{D}_0$. This size represents the whole distribution of physical sizes of cloud particles D. Sizes of individual particles D are defined in a manner as they are usually reported using aircraft in situ two-dimensional probes (e.g., the mean of maximum chord lengths measured in the parallel and perpendicular directions relative to the probe photodiode array). The characteristic size often used in modeling is the effective size D_e, which is sometimes defined as the ratio of the third and second moments of the PSD. For PSDs, which can be modeled as gamma functions of the order n,

$$D_e \approx (n + 3)(3.67 + n)^{-1}D_0. \qquad (11)$$

For water clouds, the effective diameter of cloud droplets can also be expressed as

$$D_e = 1.5\mathrm{LWC}(\rho_w P_t)^{-1}, \qquad (12)$$

where P_t is the total projected area of drops in a volume

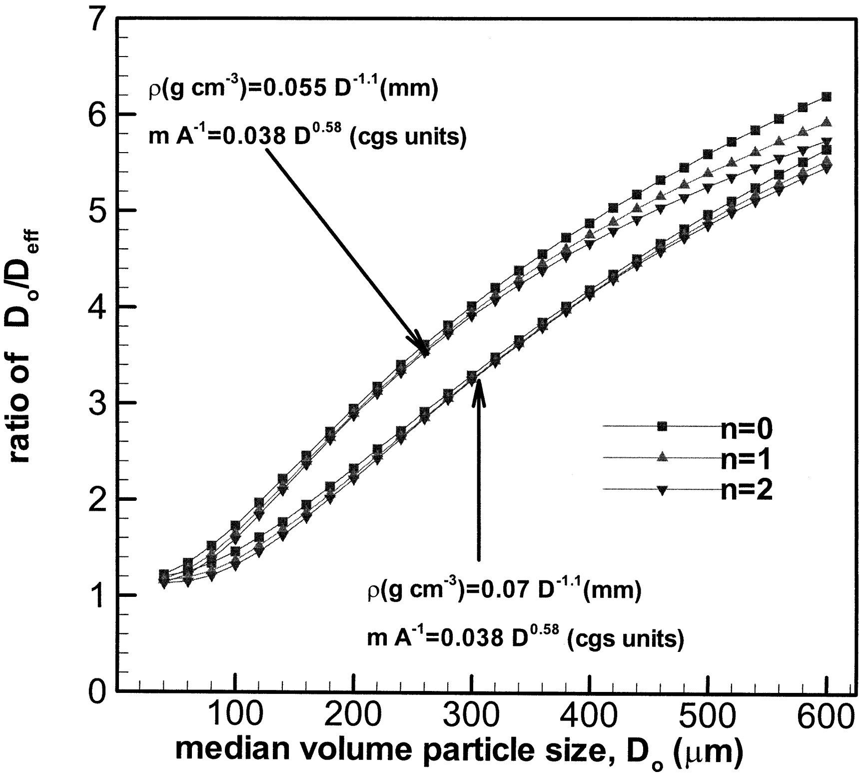

FIG. 8. The ratio of the median volume cloud particle sizes (D_0) and effective particle sizes ($D_{\rm eff}$) defined by (13) for different orders of the gamma function PSD (n).

unit, ρ_w is the density of water, and LWC is liquid water content.

Analogous to (12), it has been suggested (e.g., Mitchell 2002) that it is appropriate to use in models the effective size of ice cloud particles defined as

$$D_{\rm eff} = 1.5{\rm IWC}(\rho_i P_t)^{-1}, \qquad (13)$$

where ρ_i is the solid ice density. Unlike for water clouds, the definitions of the effective particle size as the ratio the third and second moments of the PSD (in terms of actual physical particle sizes) and as given by (13) are not equivalent because of changing bulk density and particle nonsphericity (i.e., $D_e \neq D_{\rm eff}$). Figure 8 shows the ratio r_D that defines the correspondence between D_0 and the effective size defined by (13) for $n = 1$, 2, and 3:

$$D_0 = r_D D_{\rm eff}. \qquad (14)$$

The bulk density and the cross-sectional area approximations strongly influence this ratio. Results presented in Fig. 8 assume that these approximations are described by the ρ–D and m–A–D relations discussed in section 2. For very small values of D_0, this ratio approaches to $r_D = (n + 3.67)(3 + n)^{-1}$, which corresponds to solid ice spheres. For larger D_0; $D_{\rm eff}$ is progressively smaller than D_0 (for $D_0 \approx 600$ μm, corresponding values of $D_{\rm eff}$ are about 110–120 μm).

A convenient relation between D_0 and $D_{\rm eff}$ can also be obtained using (1) and the general equation for the radar reflectivity from Matrosov et al. (2002):

$$Z_e\ ({\rm mm}^6\ {\rm m}^{-3}) = G{\rm IWC}\ ({\rm g\ cm}^{-3}) \times D_0^3 \qquad (15)$$

(where $G \approx 8 \times 10^{-5} D_0^{-1.1}$ for $D_0 > 50$ μm).

Combining (1), (15), and the definition (13) results in

$$D_{\rm eff} = 3X\rho_i^{-1}G^{-1}D_0. \qquad (16)$$

Using the approximations for X and G, which correspond to the assumptions made in this study, one can get $D_{\rm eff} \approx 18\ D_0^{0.3}$ (for $D_0 \gtrsim 75$ μm). A correspondence between the two different expressions for the radar reflectivity [i.e., (1) and (15)] is discussed in the appendix.

7. Conclusions

It has been demonstrated that the measurements taken with a ground-based vertically pointed 8-mm-wavelength cloud radar can be effectively used for retrieving optical properties of ice clouds. The suggested remote sensing method uses measurements of the Doppler velocity and radar reflectivity to derive vertical profiles of the cloud extinction coefficient, α. Vertically integrating the extinction coefficient provides estimates of cloud visible optical thickness τ, which is one of the fundamental parameters describing the radiative impact of clouds. This method is applicable to multilayer ice clouds with no strong vertical air motions. The ability of the suggested method to handle multilayer cloud situations gives it an important advantage over other existing ground-based and satellite approaches. Potentially, this approach can provide the extinction/optical thickness of the ice component in mixed-phase clouds if ice particles dominate radar moments. However, the applicability of this approach to mixed-phase clouds needs to be investigated in more detail.

The radar approach allows direct estimates of cloud extinction and optical thickness from measurements, and its results can be used as model input, bypassing some intermediate model parameterizations that relate optical and microphysical cloud properties. Simple theoretical considerations suggest that the uncertainty of the extinction coefficient retrievals could exceed a factor of 2, though the accuracy of the optical thickness estimates is generally better because of some partial error cancellations as a result of the vertical integration.

The accuracy of the radar-based optical thickness retrievals was also estimated by comparing collocated radar- and radiometer-derived values of τ. Absorption optical thickness τ_a from radiometer measurements was derived using downwelling brightness temperatures in the thermal infrared band. Estimates of the water vapor amount from the dual-channel microwave radiometer were used to account for the intervening atmospheric layer. Scattering and nonisothermic effects were also accounted for when deriving τ_a from radiometer measurements. Visible optical thicknesses from radiometer data were estimated assuming $\tau = 2\tau_a$. These radiometer-based optical thicknesses provided the robust dataset for evaluating radar-based retrievals.

Comparisons of radar-based and radiometer-based optical thickness retrievals were performed using the yearlong dataset obtained the SHEBA experiment. The radar and radiometers were essentially collocated and vertically pointed. The relative standard deviation (RSD) of radar retrievals of τ with respect to the radiometric ones was 77%, with a negative bias of -14%. One possible explanation for this negative bias is an inherent low sensitivity of radar measurements to very small particles that still may contribute to the cloud optical thickness. Uncertainties in the coefficients of the adopted m–A–D and the particle size–density relations can also contribute to this bias because the extinction retrieval results depend on these relations. Judging from the overall relatively small bias between radar- and radiometer-derived optical thicknesses, the adopted relations seem to be appropriate in describing Arctic ice cloud properties. More studies are needed to establish how well this approach for extinction retrievals and the particular m–A–D and the particle size–density relations assumed here will work under different conditions.

Taking the RSD value as a measure of uncertainty for the radar-based retrievals of cloud optical thickness results in about factor of 2 (i.e., $+100\%$, -50%) accuracy of the radar retrievals of τ. Given a rather wide range of applicability of this radar-based method, this should be considered as a positive result for a nonoptical instrument, such as radar. Similar accuracy of the optical thickness estimates can also be achieved by first deriving the IWP and layer mean cloud particle characteristic size $\overline{D}_c$ from radar-only measurements and then relating them to the cloud optical thickness using a τ–IWP–$\overline{D}_c$ relation.

An average τ–IWP–$\overline{D}_c$ relation was derived earlier based on the multisensor ground-based retrievals results for one-layer ice clouds. It was shown that this relation is in good agreement with the radar-only cloud retrievals. Comparisons with in situ cloud measurements taken during the cloud particle replicator ascents and aircraft Lagrangian descents also indicated the validity of this relation.

It was shown that Doppler radar–derived effective particle sizes are significantly greater than effective sizes used in some models because of definition differences. The correspondence between different definitions for ice cloud particles with changing bulk density was offered.

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APPENDIX

Approximate Relations between Cloud Parameters and Radar Reflectivity

Microphysical parameters of single-layer all-ice-phase clouds from different field experiments were retrieved using the Doppler radar–radiometer method (Matrosov 1997). This method allows independent retrievals of ice water content (IWC) and particle median size D_0. The independently retrieved values of microphysical parameters were then related to measured radar reflectivity values Z_e (mm^6 m^{-3}) statistically. The corresponding mean relations were found to be

$$\text{IWC (g m}^{-3}) \approx 0.12 Z_e^{0.64} \text{ and} \tag{A1}$$

$$D_0\ (\mu\text{m}) \approx 420 Z_e^{0.19}. \tag{A2}$$

Note that (A1) is very close to the empirical relation for 35 GHz suggested by Liu and Illingworth (2000), who obtained it based on calculations using in situ particle spectra: IWC $\approx 0.1 Z_e^{0.59}$.

Equations (A1) and (A2) are consistent with the general relation for the radar reflectivity (15), because combining (A1) and (A2) would approximately result in (15). This is a rather remarkable fact because IWC and D_0 in (A1) and (A2) were obtained using multisensor retrievals, whereas (A3) contains just one measureable parameter, that is, radar reflectivity. The relation (15) reflects the basic fact that the radar reflectivity is primarily a function of two cloud parameters. Different assumptions (e.g., those about particle bulk density, habit, and shape of the particle size distribution) are incorporated in the coefficient G. The variability of Z_e due to natural changes in D_0 and IWC (i.e., several orders of magnitude) is much larger than the variability of Z_e due to uncertainties of the assumptions. This, in a way, justifies considering radar reflectivity in ice clouds as a primary function of two cloud parameters. Different pairs of cloud parameters can be considered. In this article it is the extinction coefficient α and D_0 rather than IWC and D_0. Note, however, that the three parameters (IWC, α, and D_0) are interrelated through the m–A–D assumption given by (2). Using (A2) and the general relation for Z_e rewritten in terms of α and D_0 [i.e., (1)], one can obtain the following mean relation:

$$\alpha\ (m^{-1}) \approx 0.0036 Z_e^{0.58}. \tag{A3}$$

It is clear that at least two vertical profiles of input quantities are needed to retrieve independently vertical profiles of two cloud parameters such as IWC and D_0 or α and D_0. The profile of the third parameter is found using the m–A–D relationship (m determines IWC, A determines α, and D determines D_0, given assumptions about the size distribution shape and particle habit). In this article and in Matrosov et al. (2002) the two input vertical profiles are those of Z_e and V_D. The mean relations (A1), (A2), and (A3) reflect just statistical tendencies and should not be considered as robust estimates of cloud quantities. To say the least, they should not be considered as an attempt to retrieve several unknowns from one measurement of reflectivity. The uncertainies of these relations generally increase as Z_e decreases. Given these limitations, however, such relations can still be found useful for obtaining estimates of cloud parameters when only measurements of radar reflectivity are available.

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